Subaqueous Capping and Natural Recovery: Understanding the Hydrogeologic Setting at Contaminated Sediment Sites

PURPOSE: Capping and natural recovery are in situ remedial options for contaminated sediment deposits. Capping consists of the placement of one or more layers of material over a contaminated sediment deposit, while natural recovery relies upon the ongoing processes of sedimentation. One of the purposes of capping and natural recovery is the mitigation of the dissolved contaminant flux to the overlying surface water. This technical note is focused upon one of the more critical factors that determines the contaminant flux from impacted sediment deposits, the interactions of groundwater and surface water.

The purpose of this technical note is to present (1) an overview of the hydrological processes that are fundamental to understanding interplay between the groundwater and surface-water regimes, and (2) methods that have been used to measure or estimate groundwater inflow to surface water or surface-water outflow to groundwater.

INTRODUCTION: Detailed guidance on engineering considerations for dredged material capping and in situ sediment capping can be found in Palermo et al. (1998a, b). A detailed evaluation and understanding of the hydrogeology of the site is a critical component in evaluating the acceptability of a capping proposal at a proposed capping site and a prerequisite to proper cap design.

The significance of the groundwater/surface interactions is determined by the hydrogeologic characteristics of the site. Surface-water bodies are hydraulically connected to groundwater in most types of landscapes; as a result, surface-water bodies are integral parts of groundwater flow systems (Winter et al. 1998). Even if an unsaturated zone separates a surface-water body from the groundwater system, seepage from the surface water may recharge groundwater. Because of the interchange of water between these two components of the hydrologic system, development or contamination of one commonly affects the other. Consequently, a contaminated sediment deposit on the bed of a stream or lake is a potential source of contamination to the adjacent groundwater or to the overlying surface water. The extent and significance of the resulting contamination will be determined by the physical and chemical characteristics of the contaminants and of the local hydrogeologic setting.

The movement of surface water and groundwater is controlled to a large extent by the topography and the geologic framework of an area. In addition, climate, through the effects of precipitation and evapotranspiration, controls the delivery of water to and from the earth's surface. Therefore, it is necessary to understand the effects of topography, geology, and climate on surface-water runoff and groundwater flow systems in order to understand the interaction of groundwater and surface water. Furthermore, groundwater and surface-water interactions need to be understood in order to predict the potential effectiveness of an engineered cap or natural-recovery method at a contaminated sediment site. If a contaminated sediment site overlies substantial advective transport, the inflowing groundwater may cause some of the contaminants to move through the emplaced sediment cap.

20020926 049

GROUNDWATER AND SURFACE-WATER INTERACTION: Surface-water bodies occur naturally as lakes, streams, rivers, and wetlands and also exist as man-made structures, such as reservoirs. In the subsurface, water occurs in two principal zones, the unsaturated zone and the saturated zone. In the unsaturated zone, the voids (spaces between the rock particles) contain both air and water. In contrast, voids in the saturated zone are completely filled with water. Water in the saturated zone is referred to as groundwater. Within the groundwater system, geologic materials that can transmit water at rates fast enough to supply reasonable amounts of water to wells are called aquifers. Poorly permeable geologic materials that cannot readily transmit groundwater to wells are called confining beds.

Aquifers can be either confined or unconfined. A confined aquifer is overlain by a confining bed, whereas an unconfined aquifer is not. When a well is drilled into a confined aquifer, the water level in the well rises above the top of the aquifer, indicating the aquifer is under pressure. The pressure surface in a confined aquifer is referred to as the potentiometric surface. Unconfined aquifers also have a potentiometric surface, but it is in equilibrium with atmospheric pressure. This surface is referred to as the water table.

Groundwater moves from areas of recharge to areas of discharge. Recharge is the supply of water from various sources to the groundwater system. In an unconfined aquifer, recharge occurs locally by downward movement of water that infiltrates the land surface, upward movement of water from underlying geologic materials, or lateral groundwater flow. Water is supplied to a confined aquifer at an area where the aquifer crops out or where water leaks to it from a confining bed. Discharge is the movement of groundwater to the land surface, which results in saturated soils or wetlands, or to surface-water bodies, such as streams, lakes, oceans, and wetlands.

Groundwater discharging to surface water has vertical components of flow, even if the surface-water bed has relatively uniform geology (Figure 1). Many beds of surface water are more geologically complex, resulting in highly variable distribution and rates of groundwater inflow. For example, substantial inflow commonly occurs as springs, where highly permeable geologic materials intersect beds of surface-water bodies. Springs can emerge anywhere in a surface-water bed.

POSITION OF SURFACE-WATER BODIES WITHIN GROUNDWATER FLOW SYSTEMS:
The generalized flow lines in Figure 2 start at the water table, continue through the groundwater system, and terminate at the stream or at the pumped well. In the uppermost, unconfined aquifer, flow lines near the stream can be tens to hundreds of feet in length and have corresponding travel times of days to a few years. As flow paths become longer and move through deeper parts of the groundwater system, travel times can be centuries to millennia. In general, shallow groundwater is more susceptible to contamination from human sources and activities because of its close proximity to the land surface.

The type of groundwater flow system shown in Figure 2 is among the simplest types of flow systems, in that groundwater is recharged at an upland, moves downgradient through the groundwater system, and discharges to an adjacent surface-water body. Actual flow fields can be much more complex than that shown in Figure 2. For example, flow systems of different sizes and depths can be present, and they can overlie one another, as indicated in Figure 3. In these more complex groundwater systems, local flow systems are recharged at water table highs and discharge to adjacent lowlands or surface water. As local flow systems are the most dynamic, this shallowest flow system has the
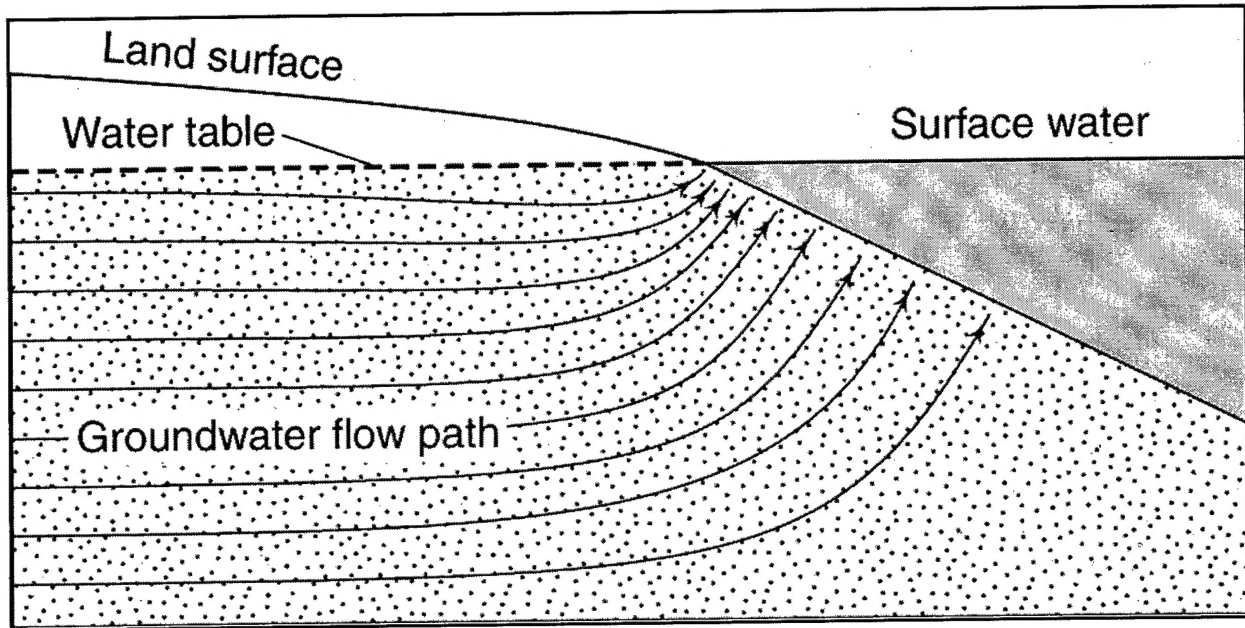
Figure 1. Groundwater inflow to surface water showing decreasing volumes of seepage with distance from shore. Modified from Pfannkuch and Winter (1984)
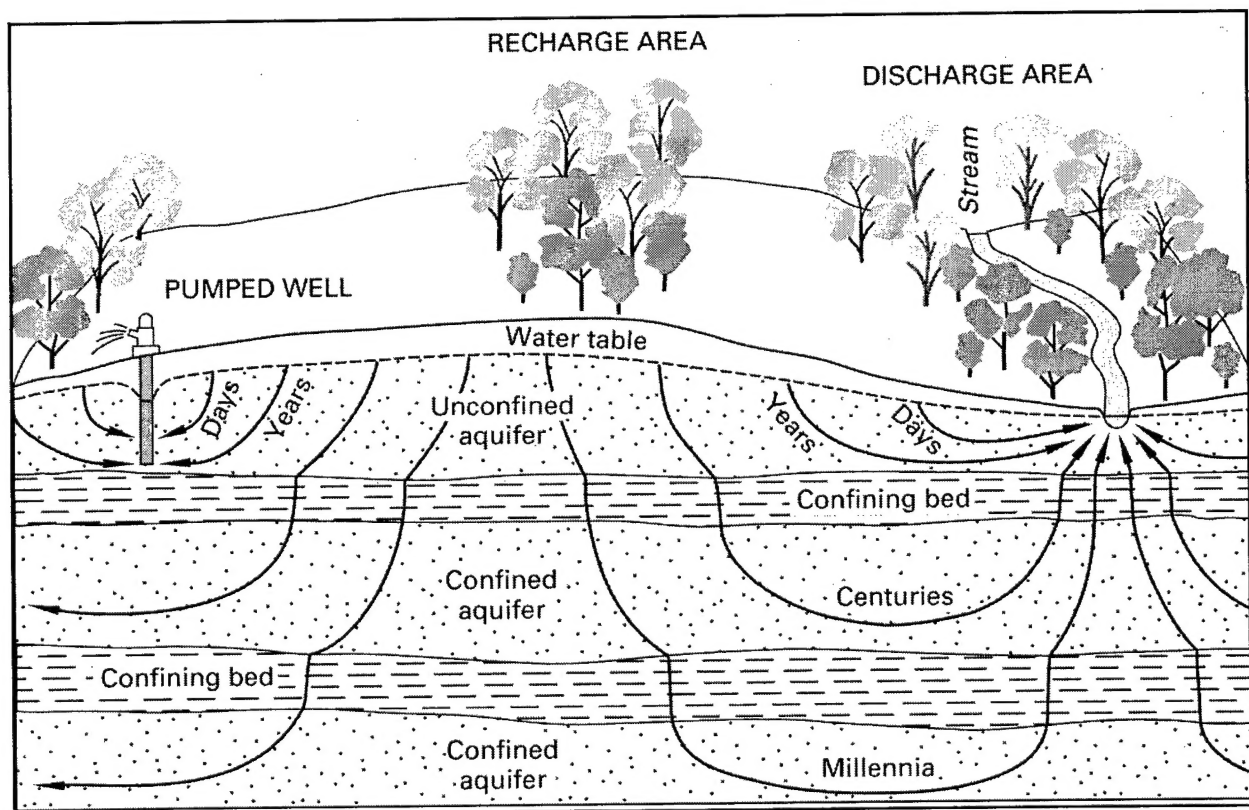

Figure 2. Schematic diagram showing groundwater flow paths having different lengths and travel times, and their relationship to surface water and a pumped well
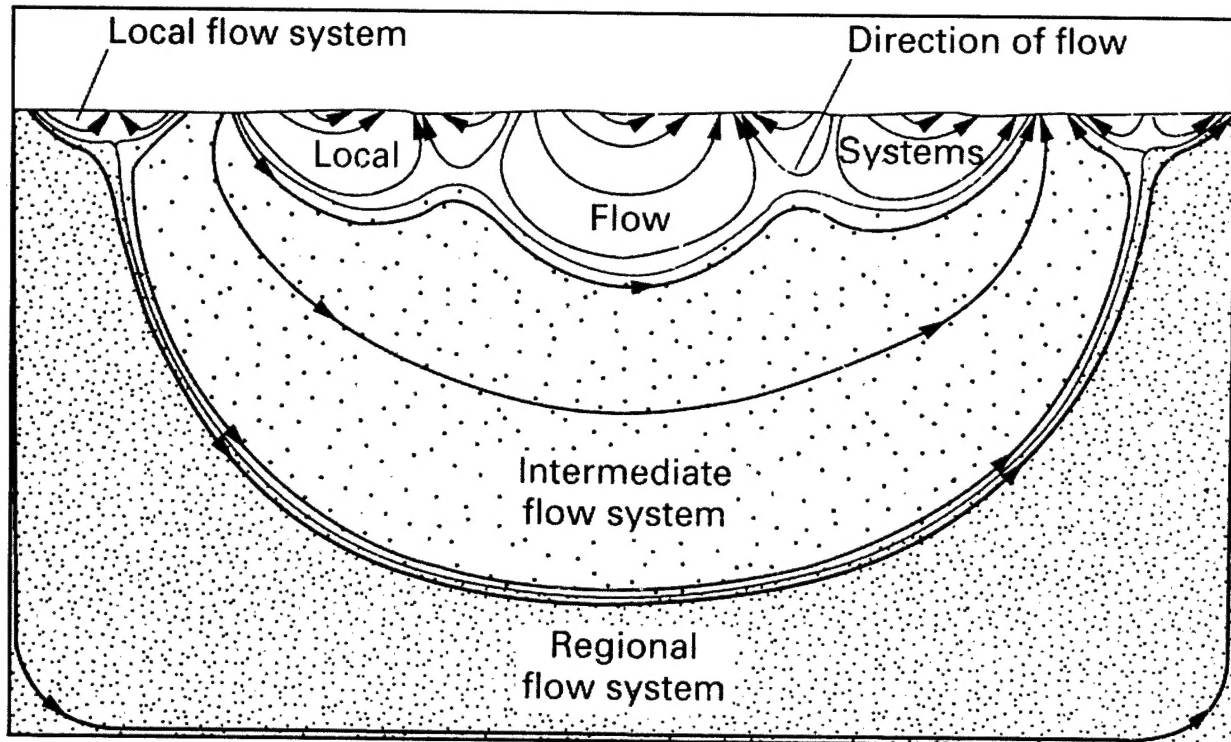
Figure 3. Schematic diagram showing groundwater flow systems of different magnitude. Modified from Toth (1963)

greatest interchange with surface water. In some areas, local flow systems can be underlain by intermediate and regional flow systems. Water in these deeper flow systems has longer flow paths, but they also eventually discharge to surface water. Surface-water bodies that receive discharge from more than one flow system receive that water through different parts of their bed. Local flow systems discharge in the part nearest shore, and larger-magnitude flow systems discharge to surface water farther offshore. Because of the different travel paths and associated travel times of water within flow paths, water discharging into surface water from different flow paths can have substantially different chemistry.

In some landscapes, surface-water bodies lie at intermediate altitudes between major recharge and discharge areas. Surface-water bodies in such settings commonly receive groundwater inflow on the upgradient side and have seepage to groundwater on the downgradient side (Figure 4). Furthermore, depending on the distribution and magnitude of recharge in the uplands, the hingeline between groundwater inflow and surface-water outflow can move back and forth across part of the surface-water bed (Winter 1986; Krabbenhoft and Webster 1995).

The above characteristics of groundwater flow systems with respect to surface water apply in a general regional sense to most landscapes. However, the detailed distribution of seepage to and from surface water is controlled by (1) the slope of the water table with respect to the flat surface of surface water, (2) small-scale geologic features in the beds of surface water, and (3) climate.
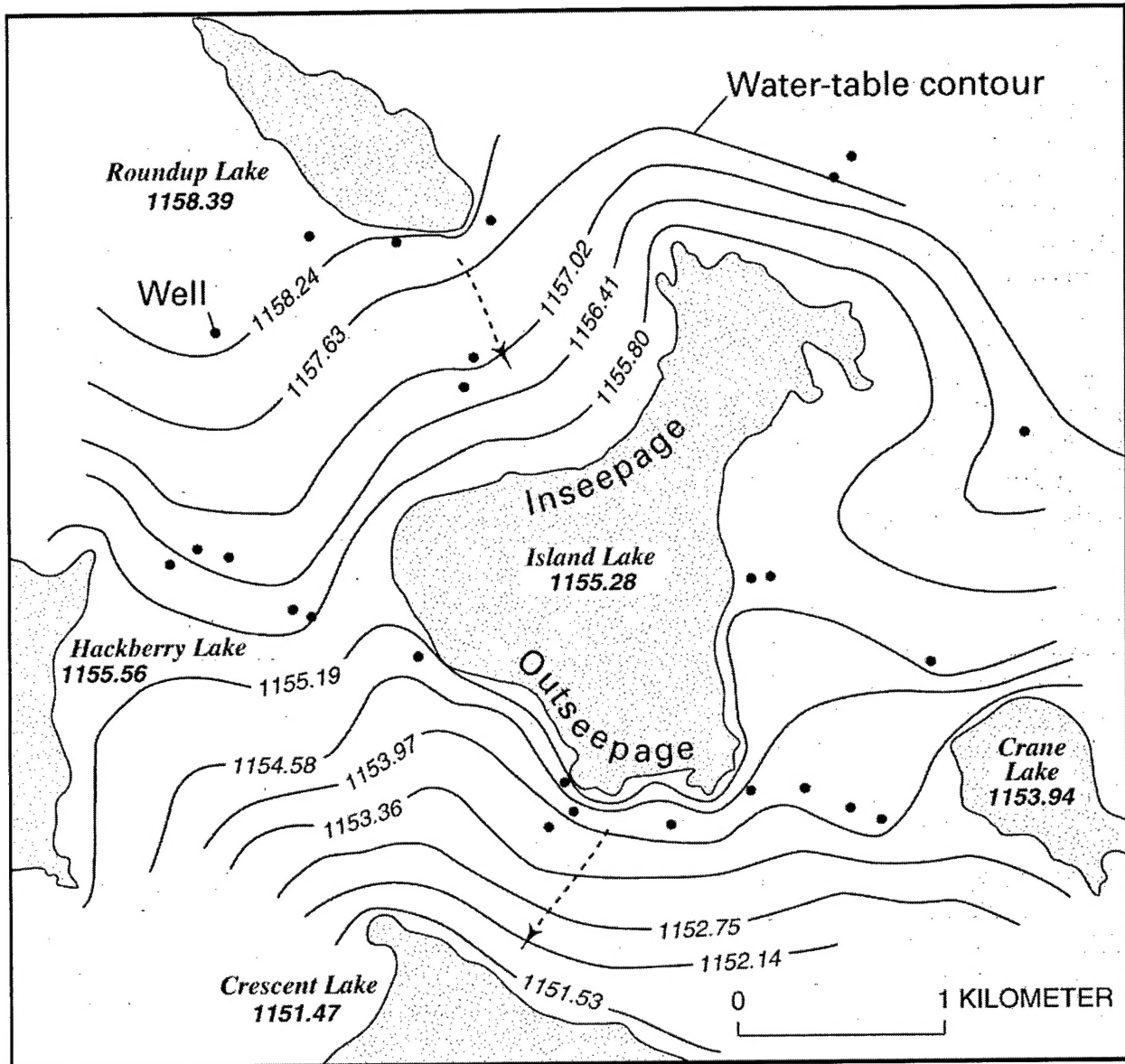
Figure 4. Map showing configuration of the water table in the vicinity of Island Lake, Nebraska. The lake receives groundwater inflow on one side and loses water to groundwater on the other. Modified from Winter (1986)

LOCAL WATER-TABLE CONFIGURATION AND GEOLOGIC CONDITIONS: Upward breaks-in-slope of the water table result in upward components of groundwater flow beneath the area of lower slope, and downward breaks-in-slope of the water table result in downward components of groundwater flow (Figure 5). These flow patterns apply to parts of many landscapes, but they are particularly relevant to the interaction of groundwater with surface water because water tables generally have a steeper slope on both the inflow and outflow sides relative to the flat surface water. The groundwater flux through a surface-water bed associated with these breaks-in-slope, whether the seepage is to or from the surface water, is not uniformly distributed. Where groundwater moves to or from a surface-water body underlain by isotropic and homogeneous porous media, the flux is greatest near the shoreline and it decreases approximately exponentially away from the
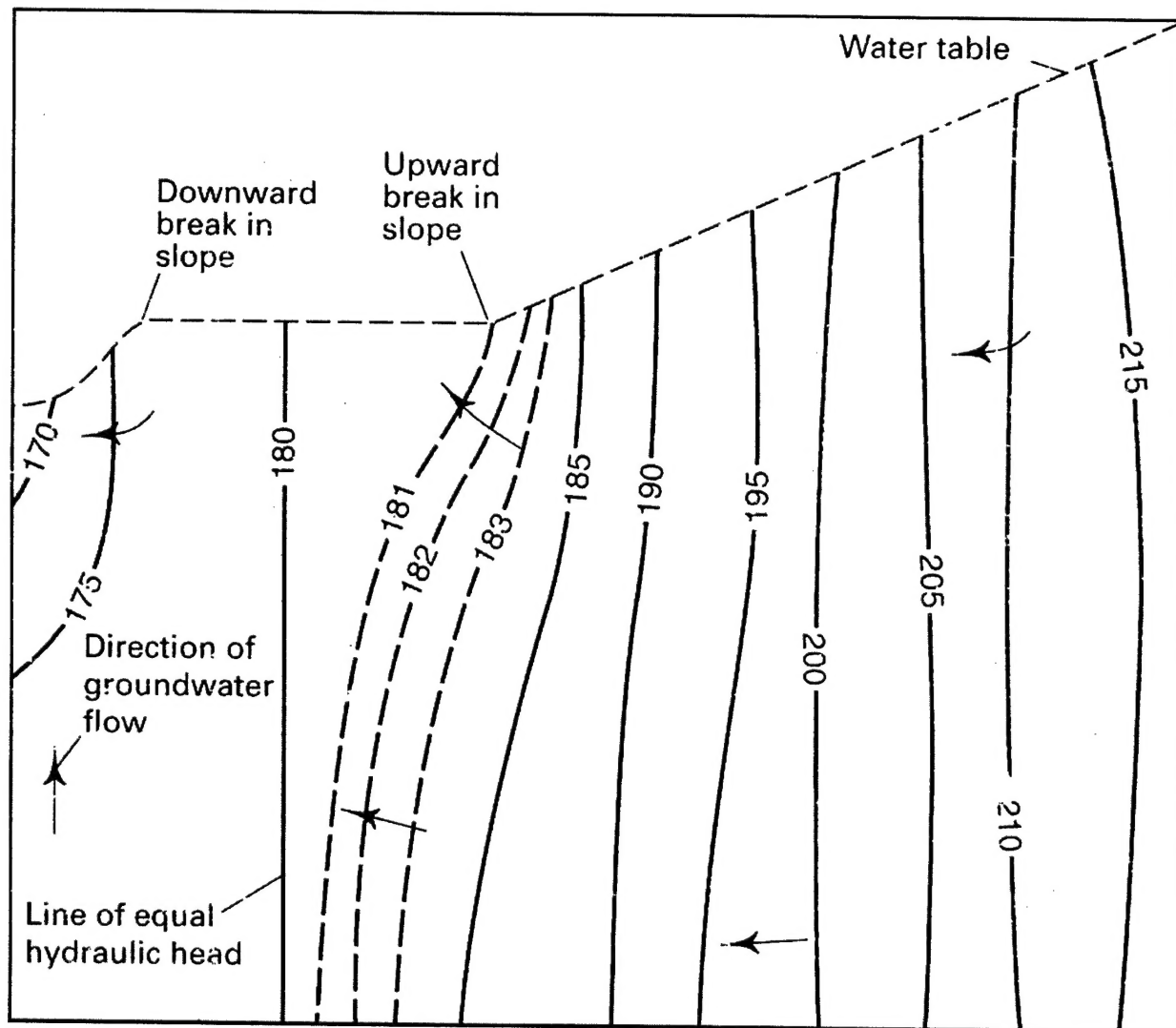
Figure 5. Schematic diagram showing the horizontal and vertical components of flow associated with breaks-in-slope of the water table

shoreline (Figure 1) (McBride and Pfannkuch 1975; Pfannkuch and Winter 1984). Anisotropy of the porous media, which is a function of the orientation of sediment particles in the geologic materials, affects this pattern of seepage by causing the width of areas of equal flux to increase with increasing anisotropy; yet, the decreasing seepage away from the shoreline remains nonlinear (Barwell and Lee 1981).

Geologic heterogeneity of surface-water beds also affects seepage patterns. For example, where groundwater heads are greater than surface-water heads, highly conductive sand beds within finer-grained porous media that intersect a surface-water bed results in subaqueous springs. In a generalized, numerical modeling study of the effect of small-scale variations in sediment type on seepage patterns, Guyonnet (1991) indicated that relatively thin, either high or low, hydraulic conductivity layers can have a substantial effect on the distribution of seepage to surface water. In a field study of the East Branch Grand Calumet River in Indiana, Duwelius (1996) found that the horizontal hydraulic conductivity of the streambed varied by five orders of magnitude and the

vertical hydraulic conductivity varied by two orders of magnitude because of the variability of streambed sediments. The complex distribution of seepage patterns caused by the heterogeneous geology of surface-water beds has been documented by field studies in many settings. The complex geologic conditions of most surface-water beds probably is the most important factor in evaluating the possible success of a sediment cap, because it results in highly variable distribution and rates of groundwater discharge to surface water.

A type of geologic setting that merits special attention is very common in areas underlain by limestone and dolomite. These areas, which are referred to as karst terrain, commonly have fractures and solution openings that become larger with time because of dissolution of the rocks. Groundwater recharge is very efficient in karst terrain because precipitation readily infiltrates through the rock openings that intersect the land surface. Water moves at greatly different rates through karst aquifers, slowly through fine fractures and pores, and rapidly through solution-enlarged fractures and conduits. The paths of water movement in karst terrain are especially unpredictable because of the many paths groundwater takes through the maze of fractures and solution openings in the rock. Seeps and springs of all sizes are characteristic features of karst terrain. In addition, the location where the springs emerge can change, depending on the spatial distribution of groundwater recharge in relation to individual precipitation events. Large spring inflows to streams in karst terrain contrast sharply with the generally more diffuse groundwater inflow characteristic of streams flowing across sand and gravel aquifers.

CLIMATIC FACTORS THAT AFFECT SEEPAGE DISTRIBUTION INTO SURFACE WATER: The water table is the most dynamic boundary of most groundwater flow systems. The configuration of the water table changes continually in response to recharge to and discharge from the groundwater system. Changes in meteorological conditions strongly affect seepage patterns in surface-water beds, especially near the shoreline. The water table commonly intersects land surface at the shoreline, resulting in no unsaturated zone at this point. Infiltrating precipitation passes rapidly through a thin unsaturated zone adjacent to the shoreline, which causes water table mounds to form quickly adjacent to the surface water (Figure 6). This process, termed focused recharge, can result in increased groundwater inflow to surface-water bodies, or it can cause inflow to surface-water bodies that normally have seepage to groundwater. Each precipitation event has the potential to cause this highly transient flow condition near shorelines as well as at depressions in uplands.

Transpiration by nearshore plants has the opposite effect of focused recharge. Again, because the water table is near land surface at edges of surface-water bodies, plant roots can penetrate into the saturated zone, allowing the plants to transpire water directly from the groundwater system. Transpiration of groundwater commonly results in a drawdown of the water table much like the effect of a pumped well (Figure 7). This highly variable daily and seasonal transpiration of groundwater may significantly reduce groundwater discharge to a surface-water body or even cause movement of surface water into the subsurface.

In many places, it is possible to measure diurnal changes in the direction of flow during seasons of active plant growth; that is, groundwater moves into the surface water during the night, and surface water moves into shallow groundwater during the day. These periodic changes in the direction of flow also can take place on longer time scales: focused recharge from precipitation predominates
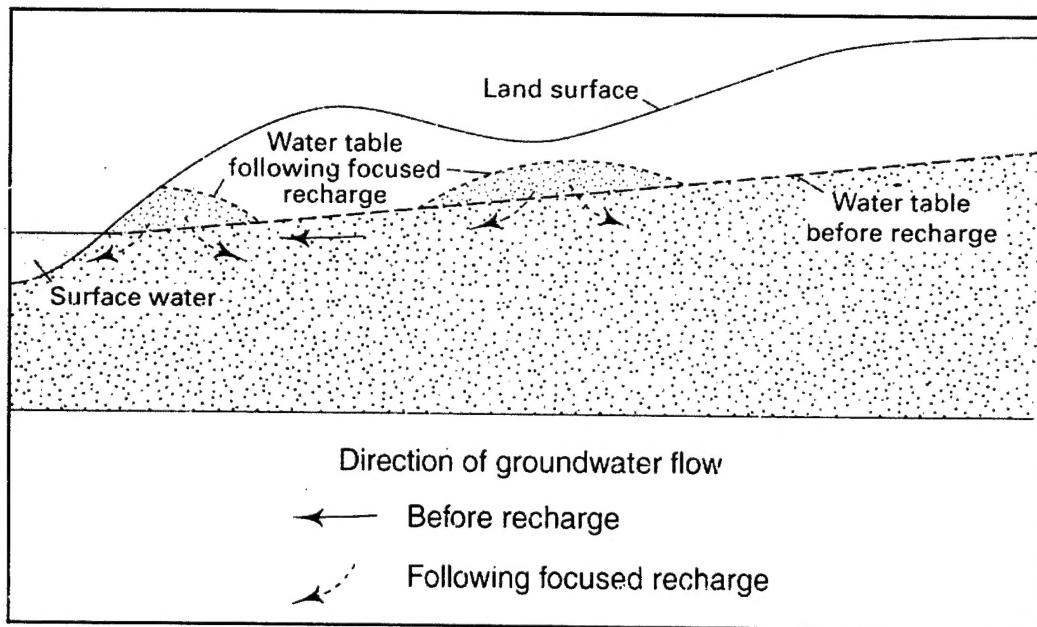
Figure 6. Schematic diagram showing transient focused recharge near the edge of a surface-water body and at a depression in the land surface
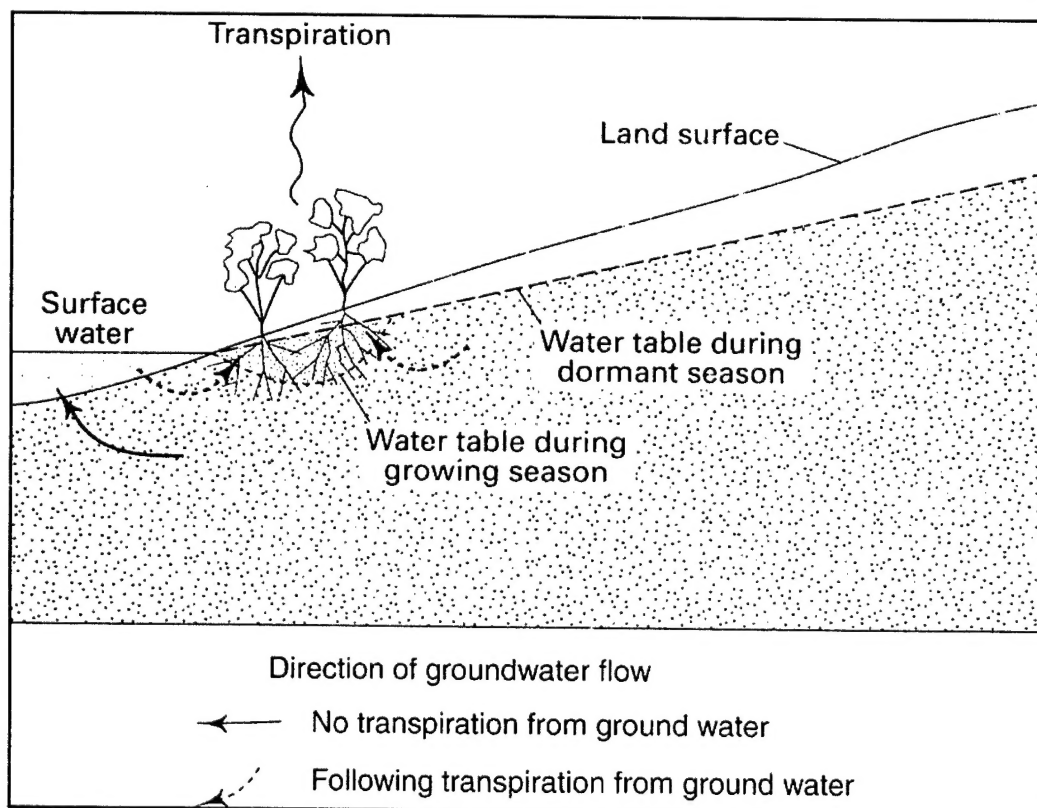

Figure 7. Schematic diagram showing transient lowering of the water table caused by transpiration directly from groundwater. The resulting cone of depression intercepts some of the groundwater that would have discharged to the surface water and can cause surface water to seep to groundwater, then to be transpired

during wet periods and drawdown by transpiration predominates during dry periods. As a result, the two processes, together with the geologic controls on seepage distribution, can cause flow conditions at the beds of surface-water bodies to be extremely variable. These processes probably affect small surface-water bodies more than large surface-water bodies because the ratio of edge length to total volume is greater for small water bodies than it is for large ones.

HYPORHEIC EXCHANGE: Streambeds and banks are unique environments because they are where groundwater that drains much of the subsurface of landscapes interacts with surface water that drains much of the surface of landscapes (Figure 8). Hyporheic exchange is the term given to the process of water and solute exchange in both directions across a streambed. The direction of seepage through the bed of streams commonly is related to abrupt changes in the slope of the streambed or to meanders in the stream channel. This process creates subsurface environments that have variable proportions of water from groundwater and surface water. Depending on the type of sediment in the streambed and banks, the variability in slope of the streambed, and the hydraulic gradients in the adjacent groundwater system, the hyporheic zone can be as much as several feet in depth and hundreds of feet in width. The dimensions of the hyporheic zone generally increase with increasing width of the stream and permeability of streambed sediments. Because of this mixing between groundwater and surface water in the hyporheic zone, the chemical and biological character of the hyporheic zone may differ markedly from adjacent surface water and groundwater.
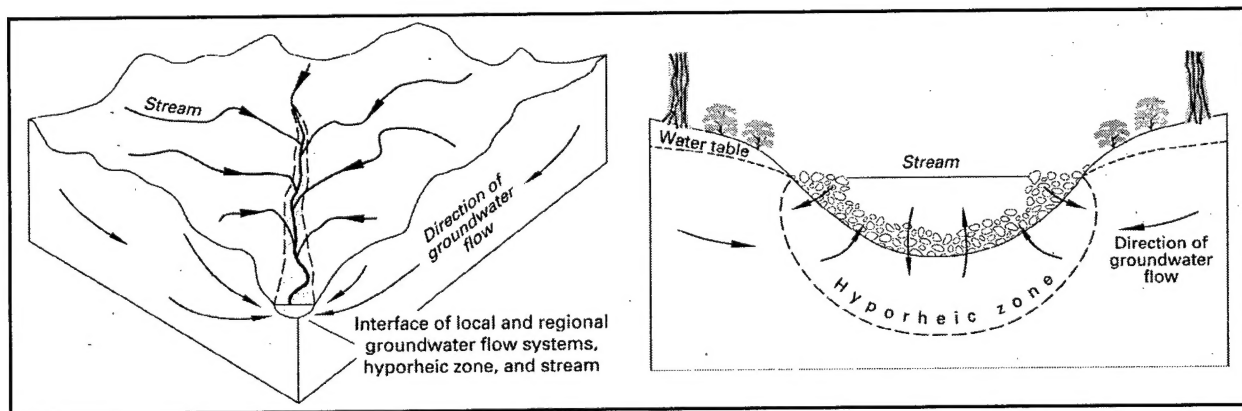

Figure 8. Schematic diagram showing the hyporheic zone as the interface between groundwater systems that move beneath large portions of the landscape and surface water that drains much of the surface of landscapes. Modified from Winter et al. (1998)

Although most research related to hyporheic-exchange processes has been done on streams, similar processes can also take place in the beds of some lakes and wetlands because of the reversals in flow. As previously discussed, this change in the direction of flow is caused by focused recharge and transpiration from groundwater. Therefore, it is necessary not only to know the relationship of surface-water to groundwater flow systems and to small-scale seepage patterns in surface-water beds, but also to be aware of hyporheic-exchange processes.

DETERMINATION OF GROUNDWATER RELATIONSHIPS WITH SURFACE WATER:
Seepage of groundwater to and from surface water ranges from slow, diffuse seepage across large areas of surface-water beds to rapid, concentrated flow at specific localities. The groundwater

contribution to, and in some cases seepage from, surface water has been determined most commonly by four methods; (1) water balance, whereby the groundwater contribution to a surface-water body is determined as the difference between all the other measured inflows to and outflows from the surface-water body, (2) hydrograph analysis, whereby the baseflow component (groundwater) of stream discharge is determined from streamflow hydrographs, (3) analytical or numerical modeling, whereby hydraulic-conductivity and hydraulic-head data from test holes and piezometers are used to calculate groundwater fluxes to and from surface water, and (4) direct measurement using seepage meters and other types of onsite measurements. For the first three methods, the estimates of groundwater interchange with a surface-water body generally apply to the overall flux across the bed without providing specific rates at specific localities. The fourth method, direct measurement, is the most common method for determining specific flux rates at specific localities in a surface-water bed. Direct measurement of seepage at specific localities is particularly useful in planning and evaluating the performance of sediment caps because interest is focused on the specific area of the contaminated sediment and the overlying cap.

Water balance. Contributions of groundwater to surface water, or losses of surface water to groundwater, can be calculated as the difference between the gains and losses of water from and to the other components of the hydrologic system, such as, precipitation, evapotranspiration, stream-flow, and overland runoff. By doing this, the value for groundwater is a net value; the actual gains and losses from and to groundwater are not determined directly. A weakness of this method is that all the errors in measuring and calculating the other components of the water balance are included in the residual value, which can result in little meaning to the term "groundwater" (Winter 1981). However, this method is included here because the calculated values for groundwater can be substantial. Despite the uncertainty in a groundwater value determined as a residual, it may give some indication of the net volume of groundwater that may be interacting with a given surface-water body, which could give an indication of the magnitude of groundwater flow. On the other hand, if groundwater inflow and outflow are both substantial, the net term may be small, resulting in an inadequate or misleading assessment of groundwater flow.

Hydrograph analysis. Hydrologists have recognized for more than a century that flow in some streams decreases at very slow rates between stormflow events, flow in others decreases very rapidly, or the streams may even go dry. It was deduced as early as the 19th century (Boussinesq 1877) that the streams with slowly decreasing flow rates were replenished by groundwater and that others had little contribution from groundwater. A number of methods have been developed for estimating the groundwater component of streamflow. Hall (1967) reviewed those that were described in the literature before the mid-1960s. More recently, Pettyjohn and Hennings (1979), the Institute of Hydrology (UK) (1980), and Rutledge (1993) developed methods whereby baseflow hydrographs are constructed from various ways of selecting and graphically connecting points of minimum flow on streamflow hydrographs. Nathan and McMahon (1990) used digital filtering technology to construct baseflow hydrographs.

Hydrograph analysis has been used to compare differences in groundwater contributions to streamflow in different physiographic settings. A study of 30 years of daily streamflow data for 54 streams in 24 landscapes comprising the contiguous United States showed that the average percentage of streamflow that was contributed by groundwater was 52 percent and the median was 55 percent (Winter, unpublished data). Groundwater contributions ranged from 14 percent in basins

underlain by silt and clay to 90 percent in basins underlain by sand and gravel. Similar studies have been done on a regional scale. For example, in a study of 114 basins in central Michigan using 3,456 station-years of daily discharge data, Holtschlag (1997) determined that the average annual groundwater contribution to streamflow ranged from 30 to 97 percent. In contrast to the relatively flat landscape of central Michigan, a similar regional study of groundwater contribution to streamflow was done in the Appalachian Mountains and Piedmont from Alabama to Pennsylvania (Rutledge and Mesko 1996). In that study, a 30-year record of daily discharge data for 89 basins indicated that the average groundwater contribution to average annual streamflow ranged from 32 to 94 percent, and the median was 67 percent.

Analytical and numerical modeling. Numerous hydrologic studies have involved calculating and (or) numerical modeling of groundwater fluxes with respect to surface water, including streams, lakes, and wetlands. The following are only a few examples of groundwater contributions to surface-water bodies determined from instrumented field sites.

In a study of the Straight River in north-central Minnesota, Stark, Armstrong, and Zwilling (1994) found that the stream discharge increased by about 26 ft^3/s over about a 15-mile reach of the river, or about 1.7 ft^3/s per mile. The stream flows across a highly permeable sand and gravel outwash plain. That groundwater is the major source of water to the river is evidenced by the fact that stream discharge decreases substantially during the summer when numerous center-pivot irrigation systems withdraw groundwater within 2 miles of the river along much of its length.

Groundwater inflow to Lake Sallie, which is located in an outwash plain in central Minnesota, was calculated using a numerical model (Larson, McBride, and Wolf 1975). The net flux rate for the lakebed as a whole was about 0.6 ft^3/s. However, the flux rates were highest at the shoreline and decreased nearly exponentially away from the shoreline. Although complementary field studies of seepage to the lake indicated that seepage rates varied seasonally, the model was of steady-state conditions; therefore, the flux rate represents an annual average.

In a study of Sparkling Lake, located in a sandy outwash plain in northern Wisconsin, Krabbenhoft et al. (1990) used a numerical model to calculate groundwater inflow and outflow rates. The average of simulations representing several seasons was 1.4×10^5 m^3/yr for inflow and 4.1×10^5 m^3/yr for outflow. The lake has a surface area of 0.81 km^2 and a volume of 8.84×10^6 m^3.

On a completely different scale, Haefeli (1972) used a numerical model to calculate groundwater flow into Lake Ontario from the Canadian side. To facilitate using the simplest flow geometry for the calculations, the flux across a vertical section at the shoreline was used because only horizontal flow needed to be considered at this line; flow systems had vertical components of flow both onshore and offshore from the shoreline. The length of shoreline considered in the calculations was about 115 miles, and the base of the cross section of flow that eventually discharged to the lake was assumed to be 300 ft below lake level. The inflow for the total cross-sectional area at the shoreline was about 64 ft^3/s.

Direct measurement. Direct measurements of groundwater discharge to and from surface water have been made in a wide variety of landscapes, from small rivers, lakes, and wetlands to the Great Lakes and the Atlantic Ocean. Measurements have been made most commonly using seepage

meters (Lee 1977) or minipiezometers (Winter, LaBaugh, and Rosenberry 1988). However, chemical methods (Cornett, Risto, and Lee 1989; Krabbenhoft et al. 1990; Jackman, Triska, and Duff 1997), and direct measurements of stream and spring discharge have also been used.

Seepage meters. Seepage meters are chambers (commonly, cutoff 55-gal drums) that are set on the bed of a surface-water body. After the chamber is allowed to settle into the sediment, a tube is inserted into an opening in the top or side of the chamber. The tube has a small bag attached at the end and a valve positioned between the chamber and the bag. The bag can be attached empty if groundwater is seeping in, or filled with a known volume of water if the direction of seepage is unknown or if surface water is seeping out. To measure the flux, the valve is opened and the change in water volume in the bag over a given period of time is a measure of flux for that period of time.

Although seepage meters are used commonly for a few measurements of groundwater flux to or from surface water, the following are a few examples of studies that made use of large numbers of meters in order to determine areal variability of seepage. In a study of Lake Sallie, (referred to previously), Lee (1972) used seepage meters that each covered 0.258 m^2 of lakebed area to measure seepage rates. Measured rates of groundwater inflow varied from 0.01 to 2.5 micrometers per second along 30 percent of the lakeshore. Groundwater inflow along an 800-m segment of shoreline amounted to 4.5×10^5 m^3/yr. Inflow rates along this segment were uniform along the shore, but they decreased exponentially away from the shoreline.

In a study of Williams Lake, located on sand and gravel ice-contact deposits in central Minnesota, Erickson (1981) used seepage meters to determine the flux to and from groundwater. Measured groundwater inflow rates were about 1×10^{-6} cm/sec at one location, and surface-water outflow rates at a number of locations varied areally from 1×10^{-6} to 14×10^{-6} cm/sec.

Shaw and Prepas (1990) used seepage meters to measure groundwater fluxes in 10 lakes underlain by glacial till in central Alberta. The meters were placed along transects that extended from the shoreline to as much as 110 m offshore. Seepage flux into the lakes ranged from 3×10^{-10} to 2×10^{-7} m/s. Groundwater contributed 49 percent of the total inflow at one of the lakes and about 10 percent of the total inflow for the others.

Asbury (1990) used seepage meters to measure seepage fluxes to and from Mirror Lake in New Hampshire. At this site, groundwater seeps into the lake from glacial till, and lake water seeps out through sand and gravel. The rates of seepage from the till to the lake ranged from less than 1 mm/day at many meter locations to about 55 mm/day at one nearshore location. Rates of seepage from the lake were far greater, ranging from near zero at some meter locations to as much as 1,000 mm/day at one nearshore location. On the outflow side of the lake, a number of meter locations had seepage rates in the hundreds of mm/day.

Belanger and Kirkner (1994) measured seepage rates of Mountain Lake, located in mantled karst terrain in Florida, using an extensive network of seepage meters. They found that the areal variability of seepage rates was much more significant than temporal variability. Individual measurements of seepage ranged from 4,533 mL/m^2/hr to the lake to 15,371 mL/m^2/hr from the lake.

With respect to water bodies of a completely different scale, Cherkauer and McBride (1988) designed and used a rugged seepage meter to conduct several studies of seepage to and from Lake Michigan. Cherkauer and Nader (1989) conducted studies of seepage along 26 transects that were located in Lake Michigan, Green Bay, Lake St. Clair, and the St. Clair River. They measured seepage rates to Green Bay that were as high as 70 mL/hr/m^2. A major finding of the study was that the seepage rates were highly variable areally, and they were controlled to a large extent by the distribution of geologic materials underlying the lake. Based on these results, they proposed a classification of seepage pattern types based on the distribution of seepage rates with distance offshore. In another study of Lake Michigan, Cherkauer and Carlson (1997) used seepage meters to define a zone of seepage from the lake caused by the drain effect of a large tunnel that was constructed near Milwaukee, WI. Seepage was induced from the lake through about a 1.3×10^7 m^2 (13 km^2) area of lakebed. Seepage rates exceeded 5 mL/hr/m^2 through about half of that area.

Chemical methods. Chemical methods to determine flux rates between groundwater and surface water have not been used as commonly as physical measurements. Two different approaches to using chemical methods are presented here. One method makes use of chemical profiles in the pore waters of surface-water sediments, and the other makes use of tracer dilution in streamflow, which is a method that commonly is used in investigations of hyporheic exchange.

Cornett, Risto, and Lee (1989) used passive, porous-membrane, pore-water collectors designed by Hesslein (1976), referred to as pore-water peepers, to determine chemical profiles in the sediments of Perch Lake in Ontario, Canada. A one-dimensional advection-diffusion model was then fit to the profiles for two nonreactive solutes, tritiated water and chloride. Tritium profiles, determined from samples collected at 2-cm intervals in the peepers, were measured where lakebed sediments are sandy and at a location farther offshore where they are organic. Calculated advective rates of groundwater flow into the lake were about 1 m/yr through the sand and about 0.1 m/yr through the organic sediments.

An example of the tracer-dilution method is provided by a study of groundwater discharge to the Shingobee River in Minnesota (Jackman, Triska, and Duff 1997). After injecting a conservative tracer such as chloride into the stream, samples were taken at five intervals along the 600-m reach of stream investigated. Groundwater inflow was then calculated for a given interval based on the dilution of the tracer concentration. The inflow rates determined apply to the entire streambed along that interval of stream reach. For four intervals within the 600-m reach of stream investigated, groundwater inflow rates varied from 0.0203 to 0.0628 L/s/m.

Seepage runs. Seepage runs have been used extensively for many years to determine groundwater inflow to streams or losses of stream water to groundwater. The method involves making stream discharge measurements at a number of locations along a reach of stream. If the measurements are made along reaches that have no tributary inflow and if they are made during a time of year that transpiration is not occurring, the gains and losses of stream water can be attributed to interactions with groundwater. As with the tracer-dilution method, the seepage rates apply to the entire streambed between the measurement locations.

Groundwater inflow to surface water as springs can also be measured directly. Springs are especially common where groundwater flow in rock fractures intersects surface-water bodies. They also are

common where small permeable deposits are present within less-permeable, finer-grained deposits. For example, springs are present around most of the perimeter of Shingobee Lake, Minnesota. The lake is underlain primarily by silty fine sand, but coarser-grained sand lenses are present within the silty fine sand. Measured discharge from the springs is as much as 2,268 L/hr.

CONCLUSION: Research and water-resource assessment studies have documented that ground-water and surface water interact in virtually all landscapes, from mountains to oceans. Movement of groundwater to surface water and of surface water to groundwater has substantial spatial variation because of heterogeneous geologic substrate and substantial temporal variation because of the effect of changing climatic conditions. This spatial and temporal variability is particularly relevant to planning and evaluating the performance of sediment caps placed over contaminated sediments and natural recovery proposals. The hydrogeology and climate of any locality having contaminated sediments need to be fully understood before the effectiveness of a sediment cap or natural recovery proposal can be determined.

POINTS OF CONTACT: For additional information, contact the author, Dr. Thomas C. Winter. United States Geological Survey, Dr. James May, (601-634-3395, *James.H.May@erdc.usace. army.mil*), or the Program Manager of the Dredging Operations and Environmental Research Program, Dr. Robert M. Engler (601-634-3624, *Robert.M.Engler@erdc.usace.army.mil*). This technical note should be cited as follows:

> Winter, T. C. (2002). "Subaqueous capping and natural recovery: Understanding the hydrogeologic setting at contaminated sites," *DOER Technical Notes Collection* (TN DOER-C26), U.S. Army Engineer Research and Development Center, Vicksburg. MS. *www.wes.army.mil/el/dots/doer*

REFERENCES:

Asbury, C. E. (1990). "The role of groundwater seepage in sediment chemistry and nutrient budgets in Mirror Lake, New Hampshire," Ph.D. diss., Cornell University, Ithaca, New York.

Barwell, V. K., and Lee, D. R. (1981). "Determination of horizontal-to-vertical hydraulic conductivity ratios from seepage measurements on lake beds," *Water Resources Research* 17, 565-570.

Belanger, T. V., and Kirkner, R. A. (1994). "Groundwater/surface water interaction in a Florida augmentation lake," *Lake and Reservoir Management* 8(2), 165-174.

Boussinesq, J. (1877). "Essa sur latheories des eaux courantes, Memoires presentes par divers savants a l'," *Academic des Sciences de l'Institut Nationale de France* 23(1).

Cherkauer, D. S., and Carlson, D. A. (1997). "Interaction of Lake Michigan with a layered aquifer stressed by drainage," *Groundwater* 35(6), 981-989.

Cherkauer, D. S., and McBride, J. M. (1988). "A remotely operated seepage meter for use in large lakes and rivers," *Groundwater* 26(2), 165-171.

Cherkauer, D. S., and Nader, D. C. (1989). "Distribution of groundwater seepage to large surface-water bodies: The effect of hydraulic heterogeneities," *Journal of Hydrology* 109, 151-165.

Cornett, R. J., Risto, B. A., and Lee, D. R. (1989). "Measuring groundwater transport through lake sediments by advection and diffusion," *Water Resources Research* 25(8), 1815-1823.

Duwelius, R. F. (1996). "Hydraulic conductivity of the streambed, East Branch Grand Calumet River, northern Lake County, Indiana," U.S. Geological Survey Water-Resources Investigations Report 96-4218.

Erickson, D. R. (1981). "A study of littoral groundwater seepage at Williams Lake, Minnesota using seepage meters and wells," M.S. thesis, University of Minnesota, Minneapolis.

Guyonnet, D. A. (1991). "Numerical modeling of effects of small-scale sedimentary variations on groundwater discharge into lakes," *Limnology and Oceanography* 36(4), 787-796.

Haefeli, C. J. (1972). "Groundwater flow into Lake Ontario from the Canadian side," Canada Department of the Environment, Inland Waters Branch Scientific Series No. 9.

Hall, F. R. (1967). "Base-flow recessions - A review," *Water Resources Research* 4(5), 973-983.

Hesslein, R. H. (1976). "An in situ sampler for close interval pore water studies," *Limnology and Oceanography* 21, 912-914.

Holtschlag, D. J. (1997). "A generalized estimate of groundwater recharge rates in the Lower Peninsula of Michigan," U.S. Geological Survey Water-Supply Paper 2437.

Institute of Hydrology. (1980). "Low flow studies," Research Report 1, Institute of Hydrology, Wallingford, United Kingdom.

Jackman, A. P., Triska, F. J., and Duff, J. H. (1997). "Hydrologic examination of groundwater discharge into the upper Shingobee River." *Hydrological and biogeochemical research in the Shingobee River headwaters area, north-central Minnesota*, T. C. Winter, ed., U.S. Geological Survey Water-Resources Investigations Report 96-4215, 137-142.

Krabbenhoft, D. P., and Webster, K. E. (1995). "Transient hydrogeological controls on the chemistry of a seepage lake," *Water Resources Research* 31(9), 2295-2305.

Krabbenhoft D. P., Bowser C. J., Anderson, M. P., and Valley J. W. (1990). "Estimating groundwater exchange with lakes 2. Calibration of a three-dimensional, solute transport model to a stable isotope plume," *Water Resources Research* 26(10), 2455-2462.

Larson, S. P., McBride, M. S., and Wolf, R. J. (1975). "Digital models of a glacial outwash aquifer in the Pearl-Sallie Lakes area, west-central Minnesota," U.S. Geological Survey Water-Resources Investigations Report 40-75.

Lee, D. R. (1972). "Septic tank nutrients in groundwater entering Lake Sallie, Minnesota," M.S. thesis, University of North Dakota, Grand Forks.

Lee, D. R. (1977). "A device for measuring seepage flux in lakes and estuaries," *Limnology and Oceanography* 22, 140-147.

McBride, M. S., and Pfannkuch, H. O. (1975). "The distribution of seepage within lakebeds," *Journal of Research of the U.S. Geological Survey* 3, 505-512.

Nathan, R. J., and McMahon, T. A. (1990). "Evaluation of automated techniques for base flow and recession analysis," *Water Resources Research* 26(7), 1465-1473.

Palermo, M. R., Clausner, J. E., Rollings, M. P., Williams, G. L., Myers, T. E., Fredette, T. J., and Randall, R. E. (1998a). "Guidance for subaqueous dredged material capping," Technical Report DOER-1, U.S. Army Engineer Waterways Experiment Station, Vicksburg, MS. *www.wes.army.mil/el/dots/doer/pdf/doer-1.pdf*

Palermo, M. R., Miller, J., Maynord, S., and Reible, D. (1998b). "Guidance for in-situ subaqueous capping of contaminated sediments," EPA 905-B96-004, U.S. Environmental Protection Agency, Great Lakes National Program Office, Chicago, IL. *www.epa.gov/glnpo/arcs/EPA-905-B94-003/EPA-905-B94-003-toc.html*

Pettyjohn, W. A. and Hennings, R. (1979). "Preliminary estimate of ground water recharge rates, related streamflow, and water quality in Ohio," Ohio State University Water Resources Research Center, Project Completion Report 552, Columbus, Ohio.

Pfannkuch, H. O., and Winter, T. C. (1984). "Effect of anisotropy and groundwater system geometry on seepage through lakebeds, 1. Analog and dimensional analysis," *Journal of Hydrology* 75, 213-237.

Rutledge, A. T. (1993). "Computer programs for describing the recession of ground-water discharge and for estimating mean ground-water recharge and discharge from streamflow records," U.S. Geological Survey Water-Resources Investigations Report 93-4121.

Rutledge, A. T., and Mesko, T. O. (1996). "Estimated hydrologic characteristics of shallow aquifer systems in the Valley and Ridge, the Blue Ridge, and the Piedmont Physiographic Provinces based on analysis of streamflow recession and baseflow," U.S. Geological Survey Professional Paper 1422-B.

Shaw, R. D., and Prepas, E. E. (1990). "Groundwater-lake interactions: II. Nearshore seepage patterns and the contribution of groundwater to lakes in central Alberta," *Journal of Hydrology* 119, 121-136.

Stark, J. R., Armstrong, D. S., and Zwilling, D. R. (1994). "Stream-aquifer interactions in the Straight River area, Becker and Hubbard Counties, Minnesota," U.S. Geological Survey Water Resources Investigations Report 94-4009.

Toth, J. (1963). "A theoretical analysis of groundwater flow in small drainage basins," *Proceedings of Hydrology Symposium No. 3*, Groundwater, Ottawa, Queen's Printer, 75-96.

Winter, T. C. (1981). "Uncertainties in estimating the water balance of lakes," *Water Resources Bulletin* 17(1), 82-115.

Winter, T. C. (1986). "Effect of groundwater recharge on configuration of the water table beneath sand dunes and on seepage in lakes in the sandhills of Nebraska," *Journal of Hydrology* 86, 221-237.

Winter, T. C., Harvey, J. W., Franke, O. L., and Alley, W. M. (1998). "Groundwater and surface water - A single resource," U.S. Geological Survey Circular 1139.

Winter, T. C., LaBaugh, J. W., and Rosenberry, D. O. (1988). "The design and use of a hydraulic potentiomanometer for direct measurement of differences in hydraulic head between groundwater and surface water," *Limnology and Oceanography* 33(5), 1209-1214.